STRATIGRAPHIC AND STRUCTURAL IMPLICATIONS OF CONODONT AND DETRITAL ZIRCON U-Pb AGES FROM METAMORPHIC ROCKS OF THE COLDFOOT TERRANE, BROOKS RANGE, ALASKA

by

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Abstract. New paleontologic and isotopic data from the Emma Creek and Marion Creek schists of the Coldfoot terrane, Arctic Alaska superterrane, central Brooks Range, suggest Devonian and possibly younger ages of deposition for their sedimentary protoliths. Conodonts from marble of the Emma Creek schist, intruded by a roughly 392 Ma orthogneiss, are late Lochkovian (early Early Devonian, between about 408 and 396 Ma) and Silurian to Devonian at two other locations. Spherical to oblong detrital zircons from quartz-mica schist of the overlying Marion Creek schist yield mostly discordant U-Pb data suggestive of provenance ages of 3.0, 2.0-1.8, and 1.5-1.4 Ga; however, several euhedral grains of zircon from Marion Creek quartz-mica schist have concordant U-Pb ages from 370 to 360 Ma. The Marion Creek schist in our study area therefore is at least 26 m.y. younger than the Emma Creek schist. The age data imply that the protolith of the Emma Creek schist is age correlative with Devonian carbonate rocks in the Hammond and North Slope terranes, whereas the Marion Creek schist is age correlative with Upper Devonian and Lower Mississippian clastic sedimentary rocks of the Endicott Group in the Endicott Mountains terrane and shale and carbonate units in the De Long Mountains and Sheenjek River terranes. Consequently, tectonic models restoring the entire Coldfoot terrane beneath partly or wholly coeval rocks of the Hammond, Endicott Mountains, De Long Mountains, and Sheenjek River terranes of the Arctic Alaska superterrane require revision. Alternative reconstructions, including restoration of the Coldfoot terrane inboard of the Endicott Mountains terrane or outboard of the De Long Mountains and Sheenjek River terranes, are plausible but require either larger amounts of shortening than previously suggested or indicate problematic facies relations.

Introduction

The Brooks Range of northern Alaska is a Mesozoic and Cenozoic contractional orogen that forms the northwesternmost part of the North American Cordillera (Figure 1). The orogen formed by telescoping of the south facing (present coordinates) Late Devonian to Jurassic passive margin of northwestern North America beneath an island arc in the Late Jurassic and Early Cretaceous [*Box*, 1985; *Mayfield et al.*, 1988; *Moore et al.*, 1994a]. The collision resulted in development of a northward-directed, thin-skinned, décollement-style fold-and-thrust belt in the northern Brooks Range and ductile deformation and high-pressure/low-temperature metamorphism in the southern Brooks Range [*Oldow et al.*, 1987; *Gottschalk*, 1990]. The style and amount of shortening in the northern (foreland) part of the orogen are well constrained by stratigraphy and abundant fossils in the deformed passive margin. In contrast, shortening in the southern, metamorphosed part of the orogen is poorly constrained due to limited stratigraphic control, sparse isotopic and fossil ages, and overprinting by multiple generations of ductile deformation and metamorphic mineral assemblages. Consequently, tectonic models and estimates of shortening for the southern Brooks Range are speculative and dependent on hypothetical stratigraphic reconstructions.

A key to reconstructing the stratigraphy, and hence the kinematic history, of the southern Brooks Range is determination of the age of the protoliths of the metamorphic rocks of the Coldfoot terrane. The Coldfoot terrane, which constitutes the schist belt of the southern Brooks Range (Figure 1 and Table 1), consists principally of polydeformed and metamorphosed sedimentary rocks, the most deformed and recrystallized metamorphic rocks of the Brooks Range. Because of their metamorphic grade, the protoliths of these rocks have been assumed Proterozoic to early Paleozoic and to constitute the oldest rocks in the Brooks Range [*Dillon*, 1989]. Published palinspastic restorations show these rocks as stratigraphic basement to allochthonous Paleozoic and Mesozoic passive-margin strata exposed in the northern Brooks Range [e.g., *Oldow et al.*, 1987; *Moore et al.*, 1994a]. However, the unconfirmed protolith ages of the Coldfoot terrane rocks leave these restorations unsubstantiated.

This paper presents newly acquired conodont and detrital zircon U-Pb ages from the Coldfoot terrane along the Dalton Highway. These data were collected as part of the Trans-Alaska Crustal Transect (TACT) project of the U.S. Geological Survey (USGS), a multidisciplinary investigation of the crustal structure of Alaska. The purpose of this study is to provide stratigraphic information that constrains structural interpretations from regional geologic map relations and integrated seismic reflection-refraction profiling of the TACT project. The data presented here are the first protolith ages from some of the most characteristic and widespread lithologies of the Coldfoot terrane. These results allow comparison of detrital zircon ages of the Coldfoot terrane with similar rocks of probable North American affinity elsewhere in the North American Cordillera and require modification of previously proposed stratigraphic and structural models for the southern Brooks Range.

Geologic Framework

The Brooks Range collisional orogenic belt resulted from southward subduction and underplating of the south facing passive margin of Arctic North America (present coordinates) beneath island arc and associated forearc and oceanic rocks of the Angayucham, and possibly the Koyukuk, terranes [*Roeder and Mull,* 1978; *Box*, 1985]. Large-magnitude contractional deformation that began in the Late Jurassic and continued through the Early Cretaceous (Neocomian) resulted in the obduction and emplacement of ophiolitic rocks of the Angayucham terrane at high structural levels in the Brooks Range [*Roeder and Mull,* 1978]. The large-magnitude contractional deformation was followed or accompanied by opening of the Canada basin to the north in the late Neocomian and regional uplift and extension along the southern margin of the Brooks Range in the middle Cretaceous [*Gottschalk and Oldow,* 1988; *Mayfield et al.*, 1988; *Grantz et al.*, 1990; *Miller and Hudson,* 1991; *Moore et al.*, 1994a]. The opening of the Canada basin may have resulted in counterclockwise rotation of all of Arctic Alaska during the Late Cretaceous [*Grantz et al.*, 1990]. Lower magnitude, thick-skinned contraction resumed in the latest Cretaceous and Cenozoic and has produced the regional structural arches and mountains of the modern Brooks Range [*O'Sullivan et al.*, 1993; *Moore et al.*, 1994b]. Estimates of total north-south shortening for the Brooks Range are in the range of 400 to 600 km [*Oldow et al.*, 1987; *Mayfield et al.*, 1988].

Prior to collisional underthrusting in the Late Jurassic and Early Cretaceous, the northwestern margin of North America consisted of a gently southward tilted passive-margin succession of Upper Devonian and Lower Mississippian fluvial-deltaic clastic rocks (Endicott Group), Mississippian and Pennsylvanian platform carbonate rocks (Lisburne Group), and Permian to Jurassic marine to nonmarine, mainly fine-grained clastic rocks (Sadlerochit Group, Shublik Formation, and Kingak Shale) [*Moore et al.*, 1994a]. Paleogeographic reconstructions indicate that the Mississippian and younger carbonate and fine-grained clastic rocks in this succession thinned and fined southward into a coeval basinal succession of predominantly siliceous shale and chert along the outer part of the passive margin. The underlying Upper Devonian and Lower Mississippian fluvio-deltaic rocks also thinned and fined southward, but the character of the southern edge of the fluvio-deltaic succession has not been determined. At least locally in the western Brooks Range, the southern margin of the fluvio-deltaic rocks is inferred to have been north of Upper Devonian and older carbonate platform rocks that are directly overlain by distal Mississippian and younger siliceous shale and chert. The nature of the underlying basement upon which the southern, distal part of the Upper Devonian to Jurassic passive-margin succession was deposited is unclear. Along the northern edge of the Upper Devonian to Jurassic passive-margin succession, basement rocks are exposed and the proximal part of the passive-margin succession rests unconformably on a deformed series of structural assemblages that include Proterozoic to Devonian platform carbonate rocks, Proterozoic quartzite and argillite, lower Paleozoic chert-argillite melange, Ordovician arc-volcanic rocks, and local Devonian granitic intrusive rocks. However, the original organization and tectonic significance of these lower Paleozoic and older structural assemblages remain uncertain.

Tectonic development of the Brooks Range orogen was described by *Box* [1985] and *Moore et al.* [1994a] and is outlined here. Subduction of basinward dipping oceanic crust began in the Middle Jurassic in the oceanic realm that lay south of the passive margin. This subduction generated an offshore accretionary prism of oceanic crustal rocks and covering siliceous sediments. During the Late Jurassic and Early Cretaceous (early Neocomian), the subducting oceanic plate consumed oceanic crustal rocks marginal to the continent and then impinged on the continental margin. The resulting underthrusting of continental crust produced large-displacement, thin-skinned thrusting and successive stacking of more distal parts of the passive-margin succession onto more proximal parts. The distal passive-margin facies now constitutes a series of allochthons assigned to either the De Long Mountains or Sheenjek River terranes (terrane terminology after *Moore* [1992]; see also Figure 1 and Table 1). Structurally beneath these rocks is the extensive Endicott Mountains terrane (allochthon) (Figure 1 and Table 1), which contains less distal parts of the passive margin than the De Long Mountains or Sheenjek River terranes. The most proximal parts of the passive margin are autochthonous or parautochthonous and comprise the upper part of the North Slope terrane, which underlies the present-day North Slope and is exposed in the northeastern Brooks Range (Figure 1 and Table 1). The Proterozoic and lower Paleozoic basement to the passive-margin succession lies beneath the North Slope and crops out in the northeastern Brooks Range, whereas basement units to the southern, more distal part of the passive margin are unknown because basement was structurally detached from the outer passive-margin succession during thrusting.

Stratigraphic reconstructions have suggested that basement units for the allochthonous, distal parts of the passive margin are in the protoliths of metamorphic rocks exposed south of the crest of the Brooks Range [e.g., *Oldow et al*., 1987; *Moore et al.*, 1994a]. The metamorphic rocks of the southern Brooks Range have been divided into two general assemblages: (1) the mostly lower grade Hammond terrane (Figure 1 and Table 1) and (2) the adjacent higher grade Coldfoot terrane to the south [*Till et al.*, 1988]. The Hammond terrane, which corresponds generally to the central belt of *Till et al.* [1988] and to the Skajit allochthon of *Oldow et al*. [1987] (Table 1), consists predominantly of quartz-rich, generally fine grained, phyllitic metasedimentary rocks, calcareous argillite, and local units of siliceous volcaniclastic and mafic igneous rocks. The Hammond is characterized by thick bodies of marble and an inhomogeneous structural style [*Till et al.*, 1988] and generally displays lower greenschist facies assemblages, although blueschist-facies rocks are locally present in the western Brooks Range [*Till et al.*, 1988]. Conodont and other fossil data indicate that the marbles of the Hammond terrane are Proterozoic(?), Cambrian, Ordovician, Silurian, and Middle Devonian [*Dumoulin and Harris*, 1987, 1995; *Dumoulin*, 1988], whereas sparse conodont and isotopic data indicate Middle and Late Devonian protolith ages for metaclastic units of the Hammond [*Aleinikoff et al*., 1993; *Dumoulin and Harris*, 1995]. Proterozoic metamorphic rocks are known in the western part of the Hammond terrane and consist of amphibolite, pelite, metaquartzite, and calc-schist [*Till*, 1989]. Rare exposures of Mississippian to Permian(?) terrigenous-clastic rocks and limestone included in the Hammond terrane by *Moore et al.* [1994a] are deformed and metamorphosed equivalents of the parautochthonous Paleozoic sequence of the North Slope terrane [*Mull and Tailleur*, 1977; *Tailleur et al*., 1977].

The Coldfoot terrane encompasses the schist belt (Table 1), a laterally continuous band 15 to 50 km wide and at least 600 km long along the southern margin of the Brooks Range (Figure 1). The Coldfoot consists of penetratively deformed and metamorphosed epicontinental clastic rocks, mafic and felsic volcanic rocks, and carbonate rocks and is intruded by Devonian granitic plutons and batholiths (now orthogneiss) derived from a continental source region [*Nelson et al*., 1993]. *Moore et al.* [1994a] assigned some of these plutons and their country rocks to the Hammond terrane, but field and age relations east of the Dalton Highway [*Aleinikoff et al.*, 1993] and north of the Ambler district (Figure 1) in the central Brooks Range [*Toro and Gans*, 1995] indicate that at least some, if not all, of these plutons should be included in the Coldfoot terrane, as reflected by the schist belt unit of *Oldow et al.* [1987]. Accordingly, we herein place the northern edge of the Coldfoot terrane farther north than shown by *Moore et al*. [1994a] to include within the Coldfoot terrane all of the major middle Paleozoic intrusive units of the southern Brooks Range.

Sedimentary and volcanic structures in the Coldfoot terrane are nearly everywhere obliterated by multiple deformations and metamorphic recrystallization, although relict granitic textures are locally preserved in the core of orthogneiss bodies. Blueschist-facies metamorphism of the Coldfoot terrane took place in the Late Jurassic and Early Cretaceous, presumably coincident with large-magnitude thrusting in the northern Brooks Range [*Armstrong et al.*, 1986]. The blueschist-facies mineral assemblages are mostly to completely overprinted by pervasive greenschist-facies assemblages that yield mid-Cretaceous K-Ar cooling ages and are thought to date the time of regional uplift [*Dusel-Bacon et al.*, 1989].

Protolith ages for rocks in the Coldfoot terrane have been reported only from the Ambler (mineral) district in the southwestern Brooks Range, where thick units of calcareous schist (Kogoluktuk schist) and mostly unfossiliferous quartz-mica schist (Anirak schist) are overlain by marble (Bornite marble) that yields Silurian to earliest Mississippian fossils [*Hitzman et al.*, 1982, 1986; *Moore et al.*, 1994a]. The Kogoluktuk schist reportedly contains rare stromatolites [*Hitzman et al.*, 1982], suggestive of a Proterozoic age. The Anirak schist contains a lens of mafic and felsic volcanic rocks, termed the Ambler sequence, which is about 1000 m thick and extends along strike for at least 90 km [*Hitzman et al.*, 1982]. A marble bed from the Ambler sequence contains poorly preserved corals that yield a questionable Late Devonian or Mississippian age [*Smith et al.*, 1978; *Moore et al.*, 1994a], and rocks correlative with the Ambler sequence about 85 km west of the Dalton Highway yielded late Early to early Late Devonian conodonts [*Dillon et al.*, 1986; A.G. Harris, USGS colln. 10622-SD, unpublished data, 1996]. U-Pb data from felsic metavolcanic rocks of the Ambler sequence suggest an Early Devonian age of 396 ± 20 Ma [*Hitzman et al.*, 1982; *Dillon et al.*, 1987]. However, the upper intercept (extrusion) age of the metavolcanic rocks was calculated using data from several different Coldfoot terrane metavolcanic rocks outside the Ambler district; therefore the ages may be inaccurate. Likewise, the fossil-bearing Bornite marble is exposed only in an isolated window south of the axis of the Coldfoot terrane, so its relation to the terrane is uncertain.

Coldfoot Terrane Along Dalton Highway

The Coldfoot terrane in the area of the Dalton Highway comprises a large structural arch nearly 30 km across (Figure 2). This arch, called the Wiseman arch by *Gottschalk* [1990], is bounded on the south by a south dipping belt of fine-grained, quartz-rich schist, phyllite, and mélange (the phyllite belt) assigned to the Slate Creek terrane by *Moore et al.* [1992] (Figures 1 and 2 and Table 1). The Slate Creek terrane marks the location of a regionally important zone of down-to-the-south faulting, along which structurally higher units have been dropped down and placed against the Coldfoot terrane on a south dipping normal fault under brittle [*Gottschalk and Oldow*, 1988] and/or ductile [*Miller and Hudson*, 1991; *Little et al.*, 1994] conditions. The nature of the northern margin of the Coldfoot terrane and its contact with the Hammond terrane is problematic in most areas [*Till et al.*, 1988]. Near the Dalton Highway, most workers have placed the northern boundary at the Wiseman (or Minnie Creek) thrust fault, a prominent south dipping fault that crosses the highway near the town of Wiseman [*Oldow et al.*, 1987; *Dillon*, 1989; *Dillon and Reifenstuhl*, 1990]. Presence of penetratively deformed schist displaying high-pressure mineral assemblages north of the Wiseman thrust caused *Moore et al.* [1991] and *Till and Moore* [1991] to place the northern limit of the Coldfoot terrane about 10 km farther north. We follow this interpretation here, setting the northern margin of the Coldfoot terrane at a north dipping fault that places the Coldfoot terrane beneath lower grade rocks of the Hammond terrane (Figure 2).

The Coldfoot terrane in the area of the Dalton Highway consists primarily of four informally named lithologic units (Figure 2): (1) Emma Creek schist (calc-schist and marble), (2) Marion Creek schist (quartz-mica schist), (3) Midnight Dome schist (quartz-rich schist, quartzite, and minor calcareous and mafic schist), and (4) Nugget Creek greenschist (mafic schist) [*Moore et al.*, this volume]. All four units are penetratively deformed, which completely obliterated all depositional features, except as noted.

The Emma Creek schist is exposed in a window into the Wiseman arch and in the hanging wall of the Wiseman thrust (Figure 2). This unit consists predominantly of brown-weathering, thinly to moderately foliated, medium-grained calcareous schist (calcite-chlorite-white mica-quartz ± albite ± epidote) and interlayered noncalcareous schist (quartz-chlorite-white mica). The calcareous schist contains moderately abundant layers and lenses of micaceous, quartz-bearing marble that are typically less than 1 m thick. Although the marbles are mostly fine to medium crystalline, some irregular zones and layers of marble are coarsely crystalline and are spotted with distinctive black calcite crystals that, in at least one place, have microscopic wormy, organic-like structures. The calcareous schist also contains prominent layers of foliated, light gray, micaceous marble up to 20 m thick. Within these marbles are abundant isoclinal folds marked by micaeous laminae. Some of the thicker marble layers can be traced more than 5 km and mark map-scale isoclinal folds within the Emma Creek schist. It is unclear if the schist was derived from a thick interbedded sequence of carbonate and clastic sedimentary rocks or whether the thicker marbles represent a single, structurally repeated layer in a calcareous clastic unit.

Near the structural core of the Wiseman arch, a body of cream-colored, intensely foliated muscovite-biotite quartz monzonite gneiss (Middle Fork Koyukuk River orthogneiss, Figure 2) is contained within calcareous schist of the Emma Creek unit. Although the contacts of the metagranitic rocks are not exposed, an original intrusive relation with the calc-schist protolith is supported by the presence of calc-silicate (tactite) rocks in contact with the granitic gneiss [*Dillon and Reifenstuhl*, 1990]. Zircons from the granitic gneiss yielded complex isotopic systematics indicating a plutonic crystallization age of about 392 Ma [*Aleinikoff et al.*, 1993]. An Early Devonian crystallization age is supported by U-Pb ages of 393± 2 and 391± 1 Ma from orthogneiss plutons located about 25 km east of the Dalton Highway in the Coldfoot terrane [*Aleinikoff et al.*, 1993].

The Marion Creek schist overlies the Emma Creek schist along a well-exposed, sharp, folded contact south of the Wiseman thrust [*Gottschalk*, 1987; *Dillon and Reifenstuhl*, 1990] (Figure 2). Although foliations in the schists are concordant across the contact, marble layers in the underlying Emma Creek schist are discordant, dipping shallowly northward relative to the contact [*Dillon and Reifenstuhl*, 1990]. This relation has led most workers to conclude that the contact is a thrust fault [*Oldow et al.*, 1987; *Dillon* 1989; *Dillon and Reifenstuhl*, 1990] (Figure 2), although an angular unconformity or extensional fault could explain the observed relations.

The Marion Creek schist consists predominantly of interlayered medium-grained quartzite and graphitic pelitic and semipelitic schist that have a total structural thickness of 7 km [*Gottschalk*, 1990]. Locally present in the unit are irregular zones and lenses of albite-rich schist and metabasite and minor calcareous schist. The lithologies are interlayered on the scale of centimeters to meters and are multiply deformed, commonly displaying complex fold interference patterns that *Dillon* [1989] described as a "knotty" texture. The dominant foliation in the schists is a transposition foliation that postdates at least one schistosity. Compositional layering parallels the older schistosity surface. Isoclinal folds and sheath folds are associated with the earlier schistosity and are refolded about younger, near isoclinal folds. The fabrics are interpreted to be sequentially formed structures developed in a single protracted contractional event with top-to-the-north sense of shear [*Gottschalk*, 1990]. However, *Little et al.* [1994], working about 35 km west of the Dalton Highway, have interpreted the younger fabric as the result of down-to-the-south shear within a zone of extensional deformation.

Both the Marion Creek pelitic schist and metabasite contain greenschist-facies assemblages (quartz-chlorite-white mica-epidote± chloritoid± garnet in pelites; albite-chlorite-green amphibole-epidote in metabasites). However, pseudomorphs after lawsonite and glaucophane in some thin sections of the Marion Creek schist and an eclogitic metabasite near Coldfoot indicate older blueschist-facies assemblages were once present in the Marion Creek unit in association with the older schistosity [*Gottschalk and Oldow*, 1988; *Gottschalk*, 1990]. The protolith of the Marion Creek schist is uncertain but is inferred to be marine, continentally derived organic-rich shales and quartzose clastic deposits with interlayered felsic and mafic volcanic rocks of Proterozoic or early Paleozoic age [*Dillon*, 1989; *Gottschalk*, 1990].

The Coldfoot terrane rocks north of the Wiseman thrust consist of a variety of metasedimentary rocks that are informally designated by *Moore et al.* [this issue] as the Midnight Dome schist and the Nugget Creek greenschist (Figure 2). The Midnight Dome schist consists of two units: (1)

structurally lower, light gray, fine-grained quartz schist (chlorite-white mica ± epidote ± garnet) and (2) structurally higher quartzite with interlayered calcareous schist, marble, and mafic schist. The Nugget Creek greenschist overlies the Midnight Dome schist and consists of dark, fine- and medium-grained mafic schist (chlorite-white mica-albite-quartz ± actinolite ± garnet ± epidote) with rare calcareous layers. Abundant pseudomorphs after blue amphibole are present in the upper part of the Midnight Dome schist and throughout the Nugget Creek greenschist. Like some marbles in the Emma Creek schist, marble in the Midnight Dome schist is locally coarse grained and contains distinctive black-calcite megacrysts. It is unclear whether the Midnight Dome schist and the Nugget Creek greenschist represent a metamorphosed stratigraphic section or, alternatively, an imbricate structural section. These units define a map-scale syncline where they have been thrust beneath the Emma Creek schist at the Wiseman thrust (Figure 2). Although there are no fossils or isotopic ages from the Midnight Dome or Nugget Creek schists, *Dillon* [1989] and *Dillon and Reifenstuhl* [1990] interpreted a Devonian protolith for the units because they correlated interlayered metavolcanic rocks 15 km to the west with Devonian metavolcanic rocks of the Ambler sequence in the Ambler district.

Sample Locations and Preparation

Detrital Zircons

About 30 kg of quartzite (Figure 2 and Table 2, sample 90TM 410A) was collected for U-Pb zircon analysis from a boulder within a pile of large blocks (Figure 3) in a quarry about 1.8 km southeast of Wiseman on the east side of the Dalton Highway. The quarry exposes rocks near the base of the Marion Creek schist, about 100 m above the inferred contact with the underlying Emma Creek schist. The precise location of the Marion Creek-Emma Creek contact in this area is uncertain because it is covered by glacial deposits. At the quarry, the Marion Creek schist consists largely of micaeous quartzite and quartz-rich mica schist. The quartzite sample is strongly foliated and contains mesoscopic isoclinal folds.

Zircons were extracted from the quartzite using standard mineral-separation methods, including Wilfley table, methylene iodide, and magnetic separator. Spherical to oblong detrital grains with frosted, pitted surfaces and smooth-faced euhedral grains with only slightly abraded tips and edges were obtained (Figure 4). A variety of rounded grains was selected to assess the range of ages of the detrital-zircon population. A large number of euhedral grains, all similar in appearance, were also analyzed to target the age of this population.

Because the zircons may have been derived from a number of provenances based on the range of color (including pink, brown, red, orange, yellow, and neutral) and morphology and results from other studies of detrital zircons from similar rocks in Alaska and Canada [*Mortenson*, 1990; *Ross and Parrish*, 1991; *Gehrels et al.*, 1995], the zircons were analyzed individually. All grains were handpicked from the (100- to 150-) mesh fraction and then mechanically abraded (method of*Aleinikoff et al.* [1990], after *Krogh* [1982]). The processed zircons were then dissolved in concentrated HF and HNO_3 in PFA Teflon capsules within Parr bombs [after *Parrish*, 1987] in an oven at about 210˚C for 3 days. The samples were spiked with a mixed ^{205}Pb-^{233}U-^{236}U tracer prior to dissolution. Pb and U were extracted from some of the samples (method after *Krogh* [1973]), whereas others were split into thirds without further chemical processing. Two thirds of each of the partitioned samples were loaded directly on filaments for Pb isotopic analysis; the remaining third, for U analysis. The resultant U-Pb isotopic data were reduced and plotted using the programs of *Ludwig* [1991a, b].

Conodonts

Samples were collected for conodont analysis from calcareous rocks of the Emma Creek schist, the Midnight Dome schist, and the Nugget Creek greenschist. None were collected from the Marion Creek schist because this unit contains few calcareous lithologies. Marbles in the Emma Creek schist, the Midnight Dome schist, and the Nuget Creek greenschist are typically highly strained, are light gray to white, and display sugary, recrystallized textures that generally indicate poor potential for conodont preservation. Two conodont samples were collected from the most prospective carbonate lithologies in the Midnight Dome schist; an additional two, from the Nugget Creek greenschist; and seven, from the Emma Creek schist. Conodonts were recovered, using laboratory techniques of *Harris and Sweet* [1989], from only four of the Emma Creek schist samples (Figure 2 and Table 2). These were all from the west wall of the Middle Fork Koyukuk River valley, 4 to 5 km west of the Dalton Highway and 5 to 9 km southwest of Wiseman. Sample 89TM 314 was collected from a 50-cm-thick, rusty-weathering, dark gray dolostone lens in Emma Creek calc-schist, less than 50 m below the contact with the overlying Marion Creek schist. Sample 88SK 104A was also collected near the contact with the Marion Creek from thinly layered, finely crystalline marble and dolostone. Sample 88SK 105A was collected from similar fissile, thinly interlayered marble that underlies a 5-m-thick massive marble bed about 200 m below sample 88SK 104A. Sample 90ATi 1C was collected from dark gray and dark brown weathering, laminated, dark gray and black, finely crystalline marble interlayered with micaceous calc-schist within about 250 m of the Emma Creek-Marion Creek.

Results

U-Pb Data From Detrital Grains in the Marion Creek Schist

U-Pb data from 27 zircon grains from the Marion Creek schist can be divided into two groups (Figure 5 and Table 3): (1) discordant data with $^{207}Pb/^{206}Pb$ ages ranging from 2833 to 566 Ma (16 grains) and (2) concordant and nearly concordant data with $^{206}Pb/^{238}U$ ages of 357 to 371 Ma (11 grains). With the exception of grain 13, all zircons of group 2 are euhedral and of various colors (Table 3). Group 1, however, also contains nine euhedral grains (1, 7, 8, 17, 20, 21, 22, 24, and 25). A common feature of many of the group 1 grains (with the exception of grains 12 and 25) is that they contain relatively low concentrations of uranium (34 to 228 ppm). In contrast, 9 of 11 group 2 zircons contain relatively high U concentrations of 559 to 1588 ppm.

On the basis of $^{207}Pb/^{206}Pb$ age, group 1 grains can be subdivided into three subgroups (Figure 5): (1) four grains (1, 8, 12, and 21) that plot on a chord with intercept ages of 3014 ± 16 and 1816 ± 14 Ma, possible times of crystallization and lead loss, respectively; (2) seven grains (3, 9, 10, 11, 15, 20, and 24) that plot near a reference chord between 1900 and 390 Ma, with degrees of discordance ranging from 6 to 34%; and (3) five discordant grains (7, 16, 17, 24, and 25) with $^{207}Pb/^{206}Pb$ ages between 1509 and 566 Ma. Because the four grains in subgroup 1 are discordant, their colinearity may be fortuitous; thus the upper intercept age may be geologically meaningless, and the $^{207}Pb/^{206}Pb$ age of each grain should be regarded as a minimum age for provenance(s). Other potential complications, such as inheritance of older radiogenic Pb and/or metamorphic overgrowths, also make determination of provenance age difficult. By carefully handpicking pristine grains, we have attempted to minimize the possibility that analyzed grains contain multiple age domains.

The seven grains in subgroup 2 yield dates similar those of to single detrital zircons from the continental-margin assemblage in the Coast Mountains of southeastern Alaska [*Gehrels et al.*, 1991] and from metasedimentary rocks of the Yukon-Tanana terrane of east central Alaska [*Mortensen*, 1990]. The array of discordant data along the 1900 to 390 Ma reference chord suggests a significant Pb loss event in the Paleozoic, perhaps related to 390 Ma plutonism [*Aleinikoff et al.*, 1993] in the Emma Creek schist. Three grains (16, 17, and 22) in subgroup 3 have $^{207}Pb/^{206}Pb$ ages of 1450, 1509, and 1330 Ma, respectively. Because these grains are only somewhat discordant (15%, 8%, and 5%, respectively), their $^{207}Pb/^{206}Pb$ ages may approximate the age of their crystalline sources. Grains 7 and 25 are 38% and 29% discordant; hence their isotopic systematics are too disturbed to estimate source ages. We conclude from this single data set that the main sources of sediment for the Marion Creek schist contained zircons that are Early

Proterozoic (2.0 to 1.8 Ga and 1.5 to 1.3 Ga, on the basis of $^{207}Pb/^{206}Pb$ ages) and Archean (3.0 Ga if the regression is appropriate) or Early Proterozoic to Archean (on the basis of $^{207}Pb/^{206}Pb$ ages) . We cannot determine if these zircons were derived directly from igneous provenances or from a sedimentary source.

Aleinikoff et al. [1984] analyzed multigrain fractions of zircons from quartzites in the Yukon-Tanana terrane, which yielded data with upper intercept ages of 2.1 to2.3 Ga. As noted by *Mortensen* [1990] and reiterated by our data set, these inferred provenance ages probably represent mixtures of many ages and, in fact, we see no evidence for 2.3 to 2.1 Ga material in our sample. *Hemming et al.* [1989] analyzed multigrain fractions of clear, euhedral, light pink zircons from the Carboniferous Nuka Formation, central and western Brooks Range, that yielded intercept ages of 2060 ± 7 and 41 ± 18 Ma (no degree of discordance given). *Nelson et al.* [1993] reported Sr and Nd isotopic data that are consistent with derivation of Devonian plutons in the Brooks Range from Early and Middle Proterozoic sources. Published and new isotopic data from samples collected throughout the Brooks Range clearly indicate a widespread Early Proterozoic (2.1 to 1.8 Ga) source area was available during Paleozoic sedimentation.

Group 2 zircons (370-360 Ma) indicate that the depositional age of the protolith of the quartzite must be younger than early Late Devonian (Figure 5 and Table 3). Although these euhedral grains may have been derived from multiple sources, they are unequivocally younger than 390-Ma granitic rocks that intruded the Emma Creek schist; thus, cannot be considered as coeval with them or with 390 Ma volcanic rocks of the Hammond terrane [*Aleinikoff et al.*, 1993].

Dillon et al. [1980] reported a composite discordia using U-Pb zircon data with an upper intercept age of 365 ± 15 Ma from several metavolcanic and metaplutonic rocks of the Hammond and Coldfoot terranes. These data include one concordant analysis of zircon from the Arrigetch Peak orthogneiss at 357 Ma (no uncertainties given). The $^{207}Pb/^{206}Pb$ ages from five Devonian volcanic rock samples of *Dillon et al.* [1980] range from 378 to 353 Ma and are 10-30% discordant. *Dillon et al.* [1987] recalculated a composite regression with intercept ages of about 396 ± 20 and about 105 Ma. This regression included some of the previously described samples. These data, however, are too imprecise to determine if some volcanic rocks of the Brooks Range might be possible sources for the euhedral zircons in our quartzite sample.

Conodont Data From the Emma Creek Schist

The Coldfoot terrane of the Brooks Range has yielded discouragingly few fossils. Consequently, the discovery of conodonts in four Emma Creek schist samples from the Coldfoot terrane in the Wiseman B-1 quadrangle, central Brooks Range, is noteworthy, particularly because one of these samples yielded a biostratigraphically diagnostic species (Table 2). The collection from sample 89TM 314 (Figure 2 and Table 2) contains four specimens of *Ozarkodina stygia* (Flajs) d morphotype (=*O. s. sorokini* Bardashev) (Figures 6d to 6f). This cosmopolitan species is known from the *Ancyrodelloides delta* Zone and the lower part of the succeeding *Pedavis pesavis* Zone of the late Lochkovian (early Early Devonian) [e.g., *Schönlaub*, 1985;*Uyeno*, 1991] (Figure 7). Ultimately, *O. stygia* may be limited to the *An. delta* Zone on the basis of ongoing revisions to the zonation of late Lochkovian pedavids and other Lochkovian conodonts [*Valenzuela-Ríos*, 1994]. The d morphotype of *O. stygia* appears to be the youngest form of this variable species [*Lane and Ormiston*, 1979; *Murphy and Matti*, 1982; *Wilson*, 1990] so that our sample indicates the upper part of the *An. delta* Zone or the lower part of the *Ped. pesavis* Zone (Figure 7). *Ozarkodina remscheidensis* (Ziegler) is the only other identifiable species from our Coldfoot terrane conodont collections (Figures 6a to 6c and Table 2). This eurytopic species is abundantly represented in rocks of middle Ludlovian (Late Silurian) to early Pragian (middle Early Devonian) age on most continents and is a common faunal associate of *O. stygia* [e.g., *Lane and Ormiston*, 1979; *Wilson*, 1990; *Uyeno*, 1991]. Conodont collections from samples 88SK 104A, 88SK 105A, and 90ATi 1C (Figure 2 and Table 2) contain poorly preserved indeterminate fragments or conservative ramiform elements that merely indicate a Silurian to Devonian (88SK 104A and 90ATi 1C) or longer age range (88SK 105A).

Conodonts from the Coldfoot terrane collections were assigned color alteration indices (CAI) of 5 to 6, and chiefly 5.5 and 6, indicating the host rock reached at least 350° to 400°C. Specimens with a CAI of 5 have abundant organic matter adhering to their surface and penetrating tectonically produced fractures. If these specimens had not been rich in organic matter, they, like the other Coldfoot terrane specimens, would have had a CAI of 5.5 or 6.

Ozarkodina stygia was widespread between 30°N and 30°S of the Early Devonian equator [*Scotese and Golonka*, 1993]. In western North America, this species ranged from north-central and east-central Alaska [*Lane and Ormiston*, 1979; this paper] and north-central Yukon Territory, Canada [*Uyeno*, 1991], southward to west-central Nevada [*Murphy and Matti*, 1982]. The species was widely distributed across southern Europe in Spain [*Valenzuela-Ríos*, 1990], Italy [*Mastandrea*, 1985], Austria [*Flajs*, 1967; *Schönlaub*, 1985], and Hungary [*Balogh and Kozur*, 1985]. Farther east, it is reported from Tajikistan [*Bardashev*, 1989] and Xinjiang Province, western China [*Wang and Zhang*, 1988]. It is also recorded from New South Wales, Australia [*Wilson*, 1990].

Ozarkodina stygia generally occurs with other pelagic realm faunas such as graptolites, ammonoids, and dacryoconarid tentaculitids [*Mastandrea*, 1985; *Schönlaub*, 1985]. The species is commonly found in flaser-bedded, nodular micritic limestone and interbedded dark shale that suggest an off-shelf, basin-margin depositional environment like that of the Selwyn basin [*Uyeno*, 1991] and Carnic Alps [*Schönlaub*, 1985]. All of the Coldfoot terrane conodont collections appear to represent a similar depositional setting. They are from mainly dark, carbonaceous, fine to very finely crystalline, rhythmically layered carbonate rocks (Table 2) suggestive of alternating hemipelagic and distal calciturbidite deposition.

Stratigraphic Implications

Data from conodonts recovered from the Emma Creek schist indicate that the protolith of the schist consists, at least in part, of Silurian(?) and Devonian carbonate-rich sedimentary rocks and includes, at least locally, lower Lower Devonian strata (408 to 396 Ma) [*Harland et al.*, 1990]. These data are in agreement with the observed, premetamorphic intrusive relation of the ~392 Ma Middle Fork Koyukuk River orthogneiss into the Emma Creek schist. Assuming a simple deformation history, the calcareous layers intruded by the orthogneiss may be somewhat older than the Lower Devonian conodont-bearing layers because the calcareous layers are situated at a structurally lower level than the conodont-bearing layers.

The detrital zircon population of the overlying Marion Creek schist includes a large component of concordant or nearly concordant grains that constrains the maximum depositional age of the sampled part of the protolith of the Marion Creek schist at 370 to 360 Ma (Late Devonian to Early Mississippian) [*Harland et al.*, 1990]. The minimum age for the unit, determined by Ar-Ar analysis [*Blythe et al.*, 1996], is constrained by the oldest ages of metamorphic mica in the study area, 130 to 120 Ma. The maximum age of the protolith of the schist is significantly younger than previous estimates for the protolith of the Marion Creek schist, which have ranged from as old as Late Archean to early Paleozoic.

A Late Devonian or younger age for the protolith of the Marion Creek schist is younger than that for the underlying Silurian(?) and Devonian Emma Creek schist but is consistent with the Marion Creek's position above the Emma Creek, assuming these units are in original stratigraphic sequence despite intense internal deformation. Although we cannot determine the nature of the contact between the Emma Creek schist and the Marion Creek schist with certainty, the location of the conodont samples 50 m below the contact and the detrital zircon sample only 100 m above suggests that at least 26 Myr is missing across the contact. This relation indicates that the contact could be tectonic, displaying younger rocks on older rocks as shown in Figure 2, or a deformed unconformity, along which a significant amount of stratigraphic section is missing.

Moore et al. [1994a] noted no unequivocal stratigraphic links between the Coldfoot terrane and characteristic parts of the Arctic Alaska superterrane, such as those in the Endicott Mountains allochthon and North Slope terrane to the north (Figure 1). However, there is no evidence of a suture separating the Coldfoot terrane from other parts of the Arctic Alaska superterrane. We therefore consider two possibilities for the affinity of the protoliths of the Coldfoot terrane: (1) the Coldfoot terrane is highly allochthonous or exotic relative to the characteristic parts of the Arctic Alaska terrane and stratigraphically unrelated to those rocks, or (2) the Coldfoot terrane is a fundamental part of the Arctic Alaska superterrane and can be linked to middle Paleozoic and possibly younger parts of the Arctic Alaska superterrane.

Nokleberg et al. [1994] distinguished the Coldfoot terrane as a tectonic body unrelated to other terranes of the Arctic Alaska superterrane because of unique stratigraphic, structural, and metamorphic characteristics potentially indicative of an origin outside of North America. Conodont species from the Emma Creek schist of the Coldfoot terrane are cosmopolitan and hence do not constrain the paleogeographic position of the Coldfoot terrane, except to indicate an origin in an offshore, equatorial region. Devonian conodonts from the Arctic Alaska terrane exclusive of the Coldfoot terrane also indicate deposition in an equatorial region [*Dumoulin and Harris*, 1995], so Coldfoot data are not inconsistent with the idea that the Coldfoot terrane originated as part of the Arctic Alaska terrane.

U-Pb data from detrital zircons in the Marion Creek schist of the Coldfoot terrane yield Archean (3.0 Ga) and Proterozoic (2.0 to 1.8 and 1.5 to 1.4 Ga) ages that are comparable to detrital-zircon suites from the miogeocline in eastern Alaska and northern British Columbia [*Gehrels et al.*, 1995]. Some of the Marion Creek schist zircon ages (1900 to 390 Ma) are also comparable to detrital zircon data from the Yukon-Tanana terrane in eastern Alaska and the northern Coast Mountains of Canada, although Late Devonian detrital zircons have not been found in metasedimentary rocks of the Yukon-Tanana terrane despite the presence of Lower to Middle Devonian metavolcanic rocks in that terrane [*Aleinikoff and Nokleberg*, 1989; *Aleinikoff et al.*, 1995]. The Coldfoot, Yukon-Tanana, and other terranes of Alaska contain Devonian plutonic rocks that *Rubin et al*. [1990] concluded constitute parts of a magmatic belt that stretched from California to Alaska along the outer miogeoclinal margin of North America. Taken together, these observations suggest that the Coldfoot terrane originated as part of northern North America, the position to which the Arctic Alaska superterrane is restored by most workers [*Lawver and Scotese*, 1990; *Moore et al.*, 1994a; *Plafker and Berg*, 1994].

If the protoliths of the metasedimentary and metaigneous rocks of the Coldfoot terrane are components of the Arctic Alaska superterrane, age and facies linkages must have existed between these tectonostratigraphic units. The Upper Devonian to Jurassic passive-margin sequence that now constitutes the Endicott Mountains, De Long Mountains, Sheenjek River, and North Slope terranes (Figure 1) is age correlative with the protolith of the Marion Creek schist and should contain stratigraphically related units. The only widespread clastic unit in the passive-margin sequence comparable to the protolith of the Marion Creek schist is the Endicott Group.

In the Endicott Mountains allochthon, the Endicott Group consists of an Upper Devonian and Lower Mississippian fluvial-deltaic sequence that forms a clastic wedge over 3 km thick [*Nilsen and Moore*, 1984] but is less than a few hundred meters thick in the De Long Mountains and Sheenjek River terranes. The Endicott Group in the Endicott Mountains terrane is conglomeratic and consists of quartz and chert-rich detritus deposited in prodelta, marginal-marine, meandering-stream, and braided-stream environments. The age of the Endicott Group (Late Devonian and Early Mississippian, ~370-360 Ma) [*Nilsen and Moore*, 1984; *Harland et al.*, 1990] is approximately equivalent to the maximum age allowed for the protolith of the Marion Creek schist by the detrital zircon data presented here. Interestingly, the Endicott Group also contains detrital zircons that are as young as 370 to 360 Ma [*Aleinikoff et al.*, 1995]. However, a <10 Myr difference exists between the crystallization ages of the euhedral zircons in both the Endicott Group and the Marion Creek schist and the depositional age of the Endicott Group determined from fossils. This variance indicates that the euhedral zircons must be either of volcanic origin or derived from a rapidly unroofed, shallowly emplaced pluton.

A volcanic origin for the euhedral component of the detrital zircons is supported by the presence of sparse lenses of mafic to felsic metavolcanic rocks in the Marion Creek schist. However, no metatuffs or felsic volcanic rocks have been recognized in the Endicott Group nor have unequivocally Late Devonian plutonic rocks been recognized in northern Alaska. Perhaps the protolith of the Marion Creek schist is the deep-marine equivalent of the Endicott Group, deposited in a region subject to volcanism that was separated by a shelfal region from the deltaic rocks of the Endicott Group. If the volcanism was submarine, deposition of tuffs in continental areas may have been suppressed by the overlying water column. Alternatively, detritus of rapidly unroofed 360 to 370 Ma granitic rocks could have been shed into separate Marion Creek and Endicott Group depositional basins from a single, unknown source region.

The strata in the Arctic Alaska terrane that are broadly similar in age to the protolith of the Emma Creek schist are carbonate rocks in the prepassive-margin part of the Arctic Alaska superterrane. Carbonate rocks compose a regionally significant part of the prepassive-margin stratigraphy of northern Alaska and are now exposed in the North Slope terrane in northeastern Alaska and in the Hammond terrane [*Dumoulin and Harris*, 1995] (Figure 1). In northeastern Alaska, a thick sequence of Proterozoic carbonate rocks (Katakturuk Formation) and a thin sequence of Proterozoic(?), Cambrian, and Ordovician carbonate rocks (Nanook Limestone) are unconformably overlain by upper Lower Devonian (Emsian) carbonate rocks (Mount Copleston Limestone) [*Blodgett et al.*, 1991]. Although highly deformed and mostly recrystallized, the thick carbonate units of the Hammond terrane (Baird Group and Skajit Limestone) typically yield Ordovician and Silurian fossils, although Proterozoic(?), Cambrian, Devonian carbonate strata are also present [*Dumoulin and Harris*, 1995]. In one locality near the Dalton highway, carbonate rocks of late Early and/or early Middle Devonian age rest on lower Paleozoic carbonate rocks, indicating the likely presence of an unconformity in the sequence [*Dumoulin and Harris*, 1995; *R. B. Blodgett*, unpublished data, 1994], comparable to that in the northeastern Brooks Range. Rocks in both areas were deposited in carbonate-basin to carbonate-platform environments. The carbonate protolith of the Emma Creek schist probably consisted of marls and other carbonate rocks that our conodont data indicate were deposited in an off-shelf, basin-margin depositional environment in the Silurian(?) and early Early Devonian. The relation among these age-correlative carbonate strata may be explained by a hypothetical facies model that places the protoliths of the Marion Creek schist in an offshore carbonate depositional environment during a time of relatively low sea level. Correlative platformal carbonate rocks would have been elevated relative to sea level and subjected to erosion, resulting in development of a disconformity coeval with offshore sedimentation. The eustatic curve for the Devonian [*Johnson and Sandberg*, 1988] shows a sea level low at the end of the Lochkovian that is consistent with this model. Alternatively, the pre-Early Devonian unconformity may record tectonic uplift related to a rifting event that affected northern Alaska in the Middle to Late Devonian [*Moore et al.*, 1994a].

If the above correlations are correct, the lithologic transition between the Emma Creek schist and the Marion Creek schist may record the change from prerift to passive-margin sedimentation. Regional stratigraphic relations in the Arctic Alaska terrane suggest that rifting occurred during the Middle Devonian and was followed by the establishment of a passive margin by the Late Devonian [*Anderson et al.*, 1994; *Moore et al.*, 1994a]. Assuming the contact between the Emma Creek schist and the Marion Creek schist is fundamentally depositional, the angular discordance between these units may have resulted from tectonism associated with the change from rift to drift tectonic processes along the outboard edge of the newly formed passive margin.

Polyphase deformational structures in the Marion Creek schist and Emma Creek schist, however, indicate that both units have undergone at least one cycle of transposition and that lithologic layering may have no relation to bedding. As a result, the upward decrease in age of the structural section may have little or no stratigraphic significance. Instead, the protoliths of the schists may have been deposited in unrelated regions, perhaps different parts of a continental slope, and juxtaposed during or prior to high-pressure metamorphism by tectonic underplating in a Jurassic to Neocomian subduction zone (*A.B. Till*, unpublished data, 1995). In this view, the observed younger on older relations may be the fortuitous result of progressive deformation in a high-strain environment.

Structural Implications

Because of the presumed early Paleozoic or older age of the Coldfoot terrane (schist belt) and its structural position relative to the Hammond terrane (Figure 8a), *Oldow et al.* [1987] restored the Coldfoot terrane to a stratigraphic position beneath the Cambrian to Upper Devonian rocks of the Hammond terrane (Figure 8b). *Moore et al.* [1994a] restored the Coldfoot terrane to a stratigraphic position beneath the Upper Devonian to Lower Cretaceous distal passive-margin deposits now exposed in the structurally high Endicott Mountains, De Long Mountains, and Sheenjek River terranes (Figure 8c). In both of these models, the Coldfoot terrane represents a part of the stratigraphic basement of northern Alaska from which stratigraphically higher rocks were detached and thrust northward. Our U-Pb data from detrital zircons in the Marion Creek schist, however, show that at least part of the Marion Creek is no older than the oldest part of the passive-margin sequence preserved in the Endicott Mountains and other allochthons and could be considerably younger. Further, our conodont data indicate that the Emma Creek unit is, at least in part, coeval with rocks of the Hammond terrane. Thus palinspastic restoration of the Coldfoot terrane to a stratigraphic position below the Hammond terrane or below the distal part of the passive-margin sequence as proposed by *Oldow et al.* [1987] and *Moore et al.* [1994a] is not tenable.

Given the known field relations and fossil and isotopic data, there are two possible alternative restorations for the Coldfoot terrane. In the first, the Coldfoot terrane could be restored to a position adjacent to rocks of the North Slope terrane and inboard of the outer part of the passive-margin sequence that now comprises the Endicott Mountains, De Long Mountains, and Sheenjek River terranes (Figure 8d). This model requires the Marion Creek schist to have a latest Devonian or earliest Mississippian depositional age correlative with basal clastic strata near the medial part of the passive-margin succession. This reconstruction is permissible because the restored basal strata decrease in age northward from Frasnian (early Late Devonian) in the distal passive-margin sequences to Visean (middle Mississippian) in the more proximal sequence of the North Slope terrane, probably reflecting regional subsidence of the passive margin [*Moore et al.*, 1994a]. The reconstruction results in juxtaposition of the 392-390 Ma orthogneiss plutons in the Coldfoot terrane with age-equivalent, compositionally similar metatuffs in the Hammond terrane [*Aleinikoff et al*., 1993].

The model, however, has the disadvantage of placing the fine-grained, Upper Devonian and(or) lowest Mississippian protolith of the Marion Creek schist in an unlikely position up depositional dip of the coeval, extensive, conglomeratic fluvial-deltaic wedge of the Endicott Group of the Endicott Mountains terrane. The model also does not explain the absence of Upper Devonian to Lower Mississippian volcanic rocks in the North Slope succession, which would be correlative with volcanic strata of the Marion Creek schist, nor does it address the present position of Mississippian and younger shales and carbonate rocks of the passive-margin sequence that are compositionally distinct from protoliths of the Marion Creek schist. Furthermore, off-shelf carbonate rocks of the Emma Creek schist would be restored between lower Paleozoic carbonate-platform rocks of the North Slope and Hammond terranes, thereby requiring a complicated shelf paleogeography.

The restoration would increase the amount of displacement of the Endicott Mountains terrane in the Early Cretaceous by a minimum of 60 km, because the Endicott Mountains terrane would be restored to a position south of the Coldfoot terrane rather than south of the Mt. Doonerak antiform, which lies at the northern margin of the Hammond terrane (Figures 1 and 8a) [*Mull et al.*, 1987]. The high-pressure metamorphic history of the Coldfoot terrane in this reconstruction would be the result of overthrusting of outboard parts of the passive margin onto the Coldfoot terrane during collisional tectonism in the Jurassic and Early Cretaceous.

Alternatively, the Coldfoot terrane could be restored to a southerly position outboard of all of the terranes of Arctic Alaska, including the distal parts of the passive-margin succession now preserved in the De Long Mountains and Sheenjek River terranes (Figure 8e). In this model, the protoliths of the Coldfoot terrane would record the transition from prerift to postrift deposition along the outboard edge of the Arctic Alaska superterrane. The high-pressure metamorphism and penetrative deformation that characterize the Coldfoot terrane could be easily explained in this model as the result of subduction to a deep structural level due to a position at the leading edge of the downgoing continental plate (the Arctic Alaska superterrane) in the Jurassic or Early Cretaceous. This reconstruction also requires less Early Cretaceous shortening in the Brooks Range than other reconstructions. Some depositional facies are also accounted for by this model; for instance, the calcareous protoliths of the Emma Creek schist of the Coldfoot terrane would be in an off-shelf position relative to the shallow water carbonates of the Hammond and North Slope terranes. Likewise, the orthogneiss plutons of the Coldfoot terrane would be restored to positions adjacent to coeval tuffs of the Hammond terrane. The Marion Creek schist, however, would be far removed from correlative clastic strata of the Endicott Group. In the restored position, the metaterrigenous clastic rocks of the Marion Creek schist would be adjacent to coeval siliceous basinal and carbonate deposits of higher allochthons of the Brooks Range (DeLong Mountains and Sheenjek River terranes). This reconstruction also calls for thrust faults to cut down section in the direction of structural transport because the Coldfoot terrane presently lies structurally beneath older rocks of the Hammond terrane.

Conclusions

New conodont and detrital zircon ages from metamorphic rocks of the Coldfoot terrane near the Dalton Highway provide constraints on age of deposition and source regions of the protoliths of metasedimentary rocks of the Coldfoot terrane. Conodonts from calcareous schist of the structurally low Emma Creek schist are Silurian to Devonian, and at one location, early Early Devonian. The conodont ages and species suggest that the carbonate protolith was deposited in an off-platform basinal environment in Early Devonian and possibly Silurian time. U-Pb ages of individual detrital zircons from a quartzite in the overlying Marion Creek schist indicate that detritus shed to this unit was derived from provenances of 3.0(?), 2.0 to 1.8, and 1.5 to 1.4 Ga, as well as of 370 to 360 Ma. Although other interpretations are possible because the data are limited to a sample from a single locality, the presence of euhedral 370 to 360-Ma zircons indicates that the protolith of the Marion Creek schist is at least in part latest Devonian or younger. Previous stratigraphic interpretations had suggested that the Marion Creek rocks, which consist of lithologies characteristic of large parts of the Coldfoot terrane, are Early Devonian or older.

The Precambrian provenance ages are similar to sources for sediment in the Yukon-Tanana terrane of east-central Alaska, the northern Coast Mountains of southeastern Alaska, and Paleozoic rocks of the northern part of the cordillera. These observations suggest that the Coldfoot terrane is not exotic to North America but rather originated as part of northern North America.

Nonetheless, the Coldfoot terrane is a complexly deformed and metamorphosed terrane that has not been investigated in detail throughout much of its 600-km length. It is presently unclear whether the age data are representative of the age of the terrane as a whole or date only parts of units that are restricted to the vicinity of the Dalton Highway. For example, about 40 km east of the Dalton Highway, *Brosgé and Reiser* [1964] showed lithologies similar to those of the Marion Creek schist as country rock for the Baby Creek batholith, which is constrained by U-Pb data to be 398 to 381 Ma [*Aleinikoff et al.*, 1993] and thus older than our protolith age for the Marion Creek schist. Proterozoic rocks have been dated at only one locality in the Coldfoot terrane in the western Brooks Range [*Karl and Aleinikoff,* 1990] but, on the basis of isotopic data, are suspected at depth elsewhere in the Coldfoot terrane [*Nelson et al.*, 1993]. For these reasons, it is likely that the Coldfoot terrane includes rocks that span Proterozoic to middle Paleozoic or younger time.

The new age data indicate that the contact between the Emma Creek schist and the Marion Creek schist, commonly mapped as a fault, places younger strata on top of older strata. Although the data allow a thrust or extensional fault contact between the two units [*Oldow et al.*, 1987; *Dillon and Reifenstuhl*, 1990], these results are consistent with a deformed, but unconformable, relation between the two units. If the contact is an unconformity, the succession records deposition that spans Devonian and possibly younger time. The Devonian is thought to be a time of rifting and formation of a south-facing, passive continental margin in Arctic Alaska [*Moore et al.*, 1994a]. An unconformity could be interpreted to reflect this tectonism. Such an interpretation must be regarded with caution, however, in view of the polyphase deformational history of and presence of transposition foliations in both the Emma Creek and Marion Creek schist units. The resulting absence of sedimentary features in the schists makes it impossible to investigate facing directions and superpositional relations that might confirm or eliminate the above stratigraphic interpretation.

Available evidence suggests that the Coldfoot terrane is an integral part of the Arctic Alaska terrane that was deformed and metamorphosed under high-pressure conditions in the Late Jurassic to Early Cretaceous. Previous structural reconstructions restored the Coldfoot terrane to stratigraphic positions beneath the Cambrian to Upper Devonian rocks of the Hammond terrane or beneath the now structurally high Upper Devonian to Lower Cretaceous distal passive-margin sequence of the Endicott Mountains, De Long Mountains, and Sheenjek River terranes. The age data reported here are inconsistent with these models and instead indicate that the Coldfoot terrane is in part coeval with at least the younger part of the Hammond terrane and the oldest part of the passive-margin sequence. The Early Devonian protolith of the Emma Creek schist may represent off-platform deposition at the time when a regional unconformity, caused either by a relative low stand of sea level or rift-related uplift, was developed in Hammond and North Slope terranes in the Middle Devonian. The Marion Creek schist, on the other hand, consists of metamorphosed quartz-rich clastic strata that have a bulk composition and apparent age similar to that of the Upper Devonian and Mississippian fluvial-deltaic strata of the Endicott Group in the Endicott Mountains terrane. Because the Endicott Group is the only widespread clastic unit of the passive-margin succession, it is the best possible equivalent for the protolith of the Marion Creek schist. Restoration of the Coldfoot terrane to structural positions correlative with, rather than stratigraphically beneath, the Upper Devonian to Jurassic passive-margin succession of the Arctic Alaska terrane requires either larger amounts of shortening during deformation in the Early Cretaceous or unlikely facies relations between major units. In either case, the present position of the stratigraphic basement for the Endicott Mountains, De Long Mountains, and Sheenjek River terranes is unknown and is an unresolved major question in the tectonic history of northern Alaska, as was pointed out by *Ellersieck and Mayfield* [1984]. Also, if the Coldfoot terrane does represent upper crustal rocks of the Arctic Alaska superterrane that are partly stratigraphically equivalent to rocks of the Endicott Mountains, De Long Mountains, and Sheenjek River terranes, it is difficult to explain why the Coldfoot terrane underwent high-pressure metamorphism, whereas these partly coeval terranes did not. Regardless of the reconstruction used, the Coldfoot terrane reached sufficiently deep structural levels to undergo high pressure, low temperature metamorphism and significant ductile strain, whereas other parts of the passive margin were underplated at relatively higher structural levels where they remained essentially unmetamorphosed.

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